Threat Assessment

Advanced Warning Operations Course

IC Severe 2

Lesson 2

Lifting Mechanisms

Warning Decision Training Branch

The title for this instructional component is “Threat Assessment.” This is the 2nd instructional component for the AWOC Severe Track. This is lesson 2 on lifting mechanisms.

Lesson 2 Learning Objectives

1. Identify types of lifting mechanisms
2. Identify the factors in evaluating lift
3. Identify the effects of orientation of line of forcing and boundary-relative steering-layer flow on convective evolution

The detection and analysis of lifting mechanisms are crucial components in short-term threat assessment. We will describe types of lifting mechanisms, how to evaluate lift , and look at the details of the lift as it relates to boundary orientation.

Lifting Mechanisms

- Evaluation of lift involves 4 factors:
  - Type
  - Scale (intensity)
  - Duration
  - Details of lift (what is being lifted, how sounding will modify)
  - Orientation (Boundary relative flow, shear vector orientation with respect to boundary, cold pool shear balance)

All lifting mechanisms are not created equal. Boundaries with the greatest persistent convergence would have the highest potential for initiating convection given all other parameters being equal. But it is difficult to define persistence based on our limitations of the observing network. Often what may seem to be convergence could be a bad wind ob, or minor fluctuations in the reported winds. Persistence adds credibility that something is real there. So here are some cautions to evaluating strength of boundaries WRT convergence:

1) Most boundaries converge air on scales too small for even mesonetworks to resolve (Crook 1996).
2) Strength of convergence doesn't always mean deep convergence. A boundary may have strong convergence down low and be capped well below the LFC.

This is a problem especially when the LFC is high and/or if there's a significant capping layer below it. Drylines are notorious for exhibiting strong surface convergence and yet shallow circulations.

One thing to remember is that when looking at the vertical continuity equation, the vertical velocity at some level depends on the integrated values of convergence/divergence below. Vertical velocity is highest at the level of non-divergence where convergence resides below and divergence above. In most cases, the deeper the convergence, the higher the level of maximum vertical velocity will occur.

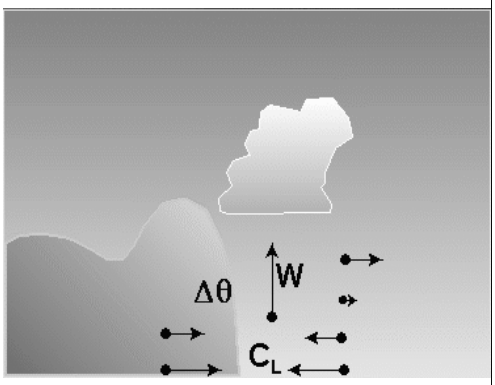
Types of Lift

- Boundaries
  - Evaluate strength by analyzing density gradient at surface and aloft (θ, T, Td, and P)
  - Evaluate convergence by analyzing moisture divergence
  - Best lift when coupled with jets

There are many types of lifting mechanisms, low-level boundaries being the most common. Depending on the scale, observable boundary attributes are location, density gradients, low-level convergence and ascending air. Density gradients produce direct mesoscale ascent, so that's why we analyze direct observables such as T, Td, q, P, and θ (actually θv is the best for density). Convergence can be estimated by calculating differential velocity across the width of the boundary (See http://www.nssl.noaa.gov/istpds/icu92/icu922web/ic922.html for more details). Ascending air, W, can be estimated by integrating values of convergence with height. Estimating one-dimensional convergence assumes a uniform airflow on either side. In other words, a front with no undulations or intersections with other boundaries. Convergence will be higher at intersections, mergers, collisions with other sources of ascent. Convergence is scale dependent. Data from METARs and mesonets will underestimate convergence in boundaries by 2 to 3 orders of magnitude. Model analysis will underestimate convergence similarly. All operational model resolutions are too poor to capture the true width of a boundary. Also, stormscale boundaries are typically unobserved by models. Typical boundaries are 3-10 km wide. Actual convergence will be occurring on these scales. Convergence will be higher at boundary intersections with other boundaries, circulations, rolls. Two dimensional convergence with boundary intersections becomes important since winds vary both along and normal to the boundary axis. Surface convergence may not always mean deep convergence/ascent . A capped atmosphere may limit the depth of even strong surface convergence. Therefore, the deeper the convergence zone relative to the LFC, the more likely initiation will occur.

Observable Boundary Attributes

- Location and movement
- Density gradient, Δθ
  - Except for troughs and HCRs
- Low-level convergence, C_L
  - Change with height
- Observing ascending air, W

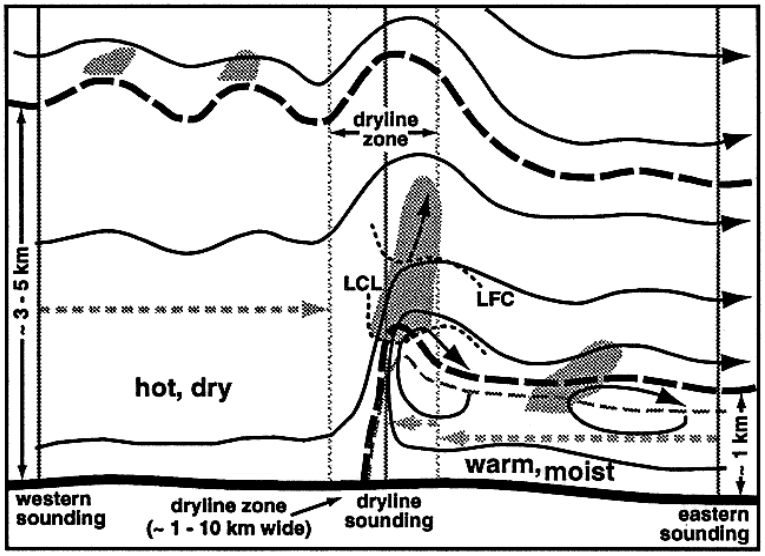


All of these are important when assessing lift

When analyzing boundaries as a potential lifting mechanism, try to evaluate the following:

1. Location and movement
2. Density gradients (which creates a thermally direct circulation and ascent on the least dense side of a boundary)
3. Low-level convergence (typically strongest near the surface)
4. Ascending air, W (even if you have convergence at the surface, it doesn't necessarily imply ascent). It has to be integrated values of convergence with height (decreasing convergence with height for example). Visual manifestation of moist W is cumulus clouds.

Lifting Along the Dryline



From Ziegler and Rasmussen (1998)

Moist boundary layer air parcels must be lifted to their LCL and LFC prior to leaving the mesoscale updraft to form deep convection. Cloud formation is predicted when the vertical mesoscale moisture flux predominates below the LCL, and deep convection is predicted if strong mesoscale lifting is deeper than the LFC.

A modification of proximity soundings to account for mesoscale lift, the across-dryline differences of environmental thermal stratification, and westerly wind shear effects can improve the diagnosis of the mesoscale dryline environment and the prediction of convective initiation at the dryline (Ziegler and Rasmussen, 1998).

Types of Lift

- Lower branch of indirect circulation
  - Enhances θ_e transport, low-level jet, convergence
- Upper divergence
  - From jet streaks, large scale ascent from Q-vector divergence
  - May be important to monitor when near-surface layer convergence is weak

In addition to boundaries, there are other lifting mechanisms like these shown here, which produce synoptic scale ascent. Most of these types of lifting can support the development of convection.

Types of Lift

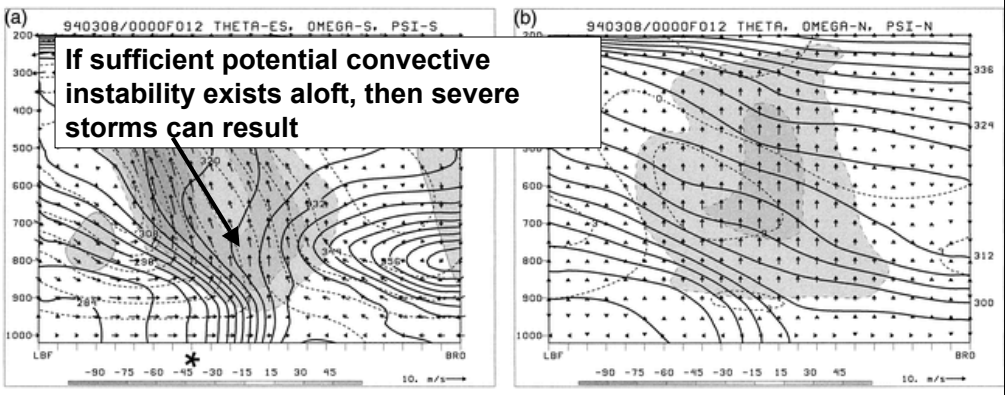
- Frontogenesis
  - Analyze incremental layers from 700 to 500 mb determined from AWIPS cross section drawn perpendicular to the temperature gradient
- Isentropic lift
  - Related to warm air advection and potential elevated convection if sufficient MUCAPE present

The diagnosis of frontogenesis will result in a diagnosis of the forcing for vertical motion on the frontal scale.
Ascent occurs on the warm side of a maximum of frontogenesis and on the cold side of a region of frontolysis (Schultz, 2001)

Elevated Convection and Frontogenesis

circulation **within** plane of cross section
(i.e., frontal circulation)

circulation **normal** to plane of cross section
(i.e., synoptic-scale circulation)



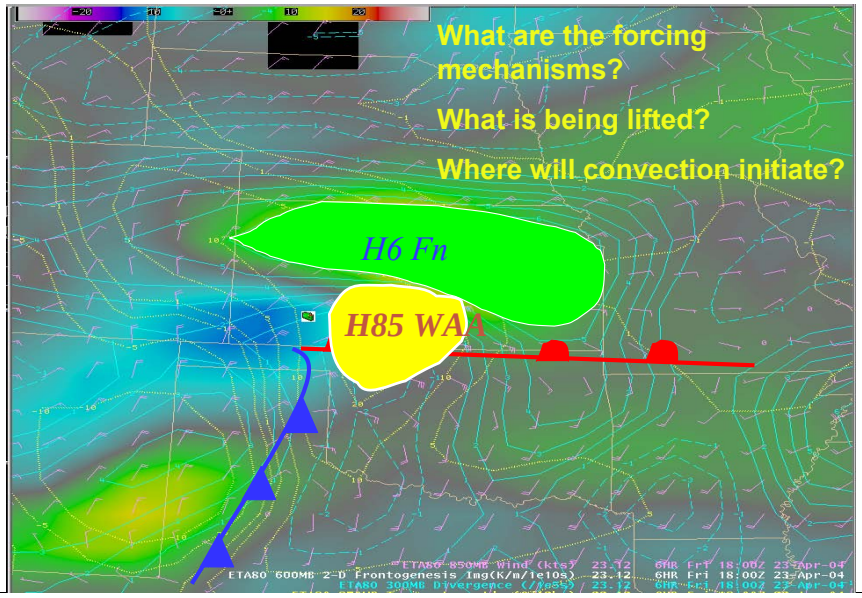
Vertical motion: shaded
Theta-Es: solid lines

Vertical motion: shaded
Theta: solid lines

From Trapp et al. (2000)

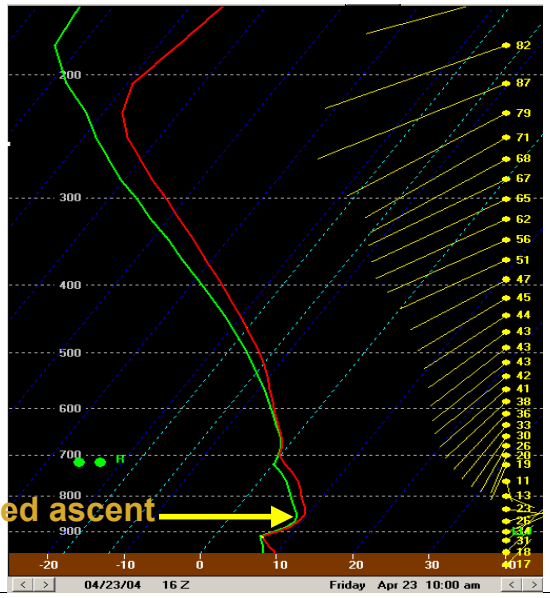
If sufficient instability is available aloft, severe elevated convection can result.

Combo of Lifting Example 4/23/04

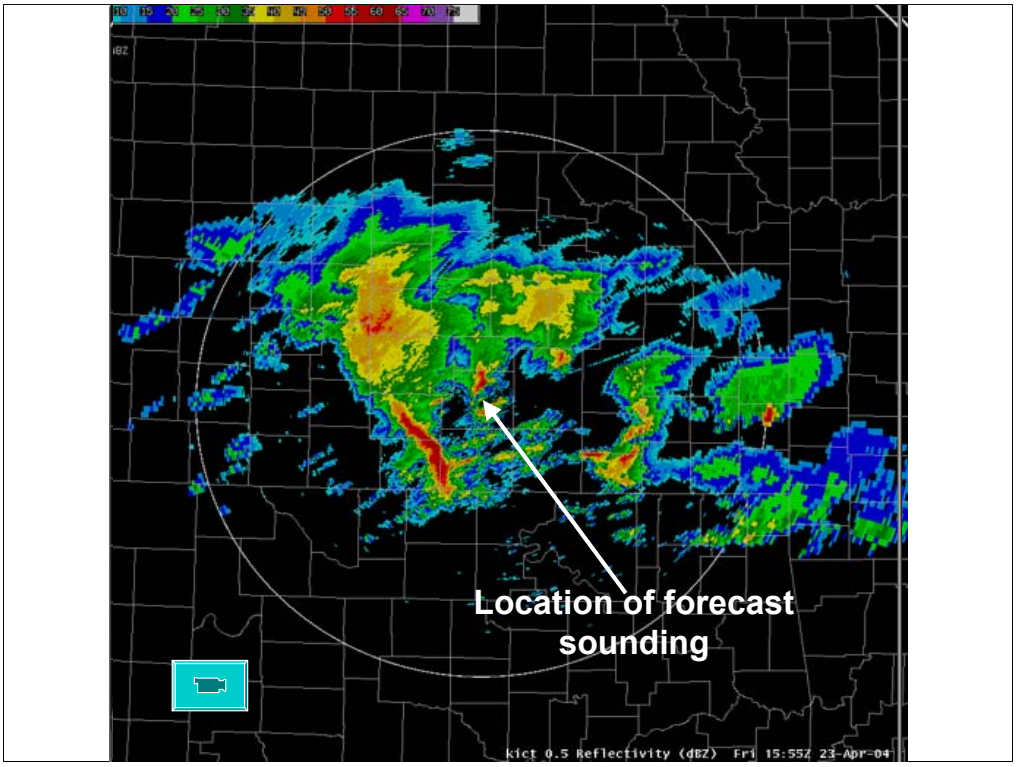


There were multiple contributions to upward motion indicated by the Eta across KS; all the convection which occurred, including the severe thunderstorm complex which developed and moved across KS during the morning hours was not surface based.

Sounding Showing Elevated Parcel Lifting Necessary



This is an Eta 4 hr forecast sounding for central KS - note the lifted parcel level (level where the parcel has the most unstable characteristics) is around 850 mb. CAPE when lifted from this level results in about 400 j/kg.



At this slide, you will want to play the flash movie showing the radar evolution. Note the convection developed into a forward propagating system – likely due to storm-relative winds from the east at low levels enhancing propagation effects.

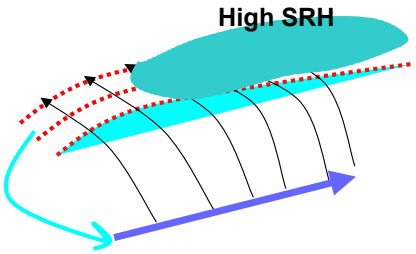
Types of Lift

- Differential heating
- Upslope winds

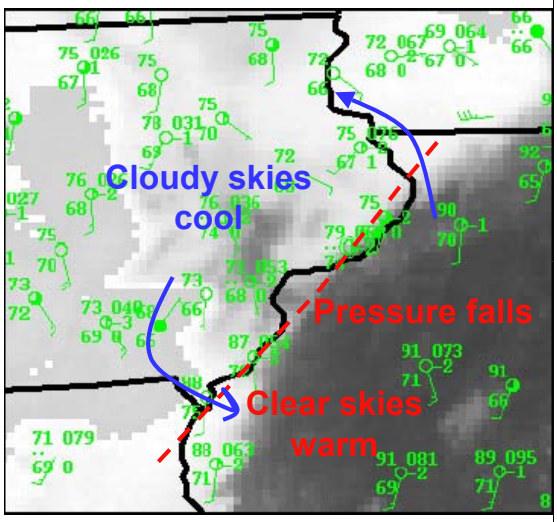
We'll take a look at lifting from differential heating.

Differential Heating

- **Over time differential ground heating will produce a solenoidal circulation**



- **Thermal gradient can also act to back low-level flow and increase near-ground SRH**



From Wolf (2002)

Pressure rises near the top of the boundary layer over the warmer air mass. This leads to divergence aloft and surface pressure falls within the warm sector. The adjacent cooler air subsequently flows toward the area of lowered pressure, and a direct circulation (warm air rising, cold air sinking) results. The increase gradient produces a solenoidal circulation, which depending on amount of CAPE/CIN and external lifting mechanisms, can help initiate severe convection.

Example at right is from 14 June 2001 (Wolf, AMS SLS conf.)

Upslope Lifting

- Terrain effects can aid boundary layer low-level jet characteristics
- Also, flow effects due to topography (see http://meted.ucar.edu/mesoprim/flowtopo/)

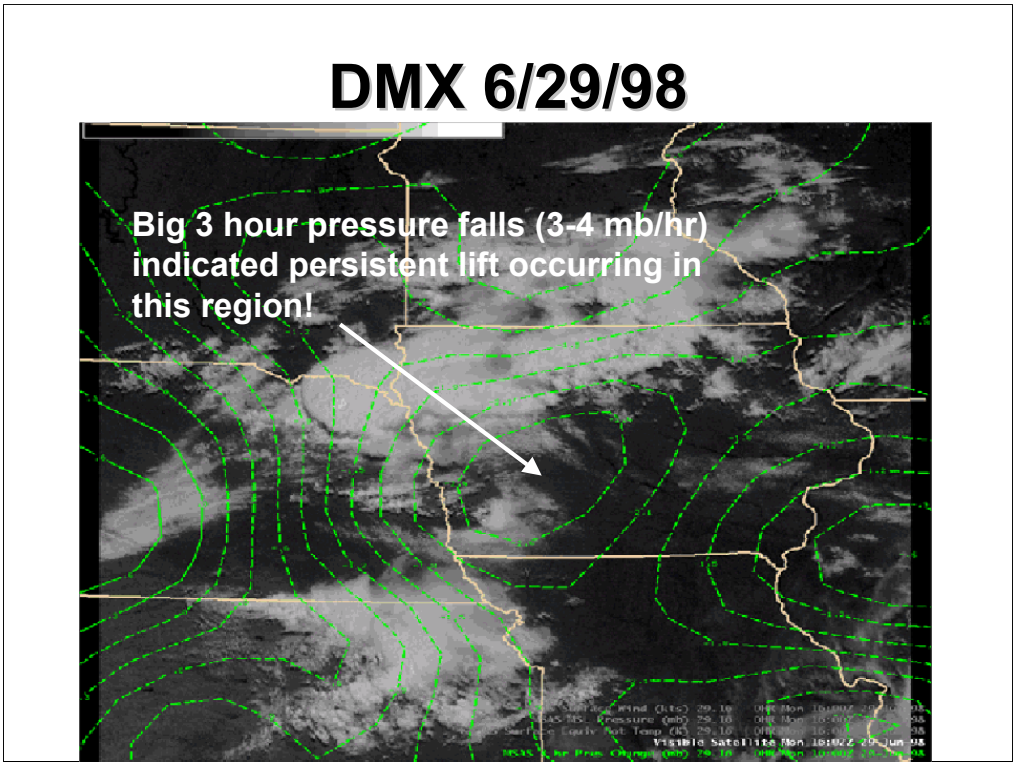
This is obviously an important consideration where there are significant changes in topography and wind flow.

Duration of Lift

- Surface pressure falls
  - 2 and 3 hour falls with continuity (often fall/rise couplets)
  - Location of fronts and movement of meso-lows by analyzing isallobaric pattern

In the DMX derecho event of 29 June 98, there were 3.6 mb/hr pressure falls in central IA for 3-4 hours prior to line intensifying and mesolow development.

One of the results was increased low level winds into the leading edge of the line, which increased vertical shear and updraft strength for storms which developed along the leading edge (some spawned tornadoes and 100 mph+ winds).



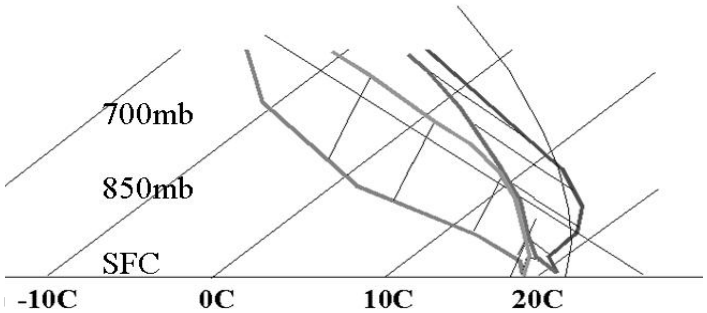
This event was analyzed in the DMX 29 June 1998 WES case.

Details of Lift

Synoptic scale lifting

With this unrealistically strong amount of lift for a 12-hour period, you saturate low-levels, and weaken cap sufficiently for moist convection.

In reality, vertical motion is substantially weaker, and varies in intensity with height, so sounding does not modify the same amount throughout.

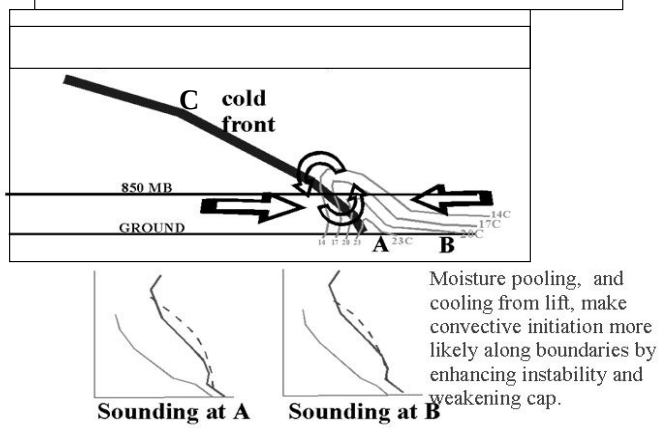


Graphics courtesy of Pete Wolf(ICT)

This is for synoptic scale layer lifting, which over a period time (0-12 hours) can act to significantly destabilize the atmospheric profile. This development of instability due to vertical motion on the large scale can contribute to convective development, but the instability alone provides inadequate mesoscale lifting for severe thunderstorms.

Details of Lift

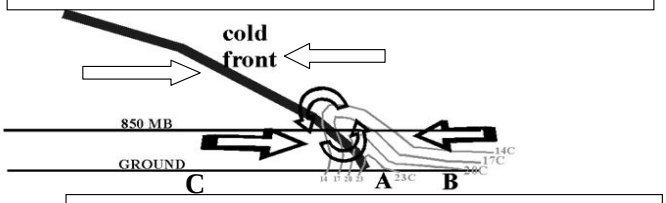
Determining what layer is being lifted



Use a surface based parcel along boundaries, a mixed layer parcel in warm sector, and a most unstable parcel north of a parcel (when instability is not rooted in boundary layer). Graphics courtesy of Pete Wolf (ICT).

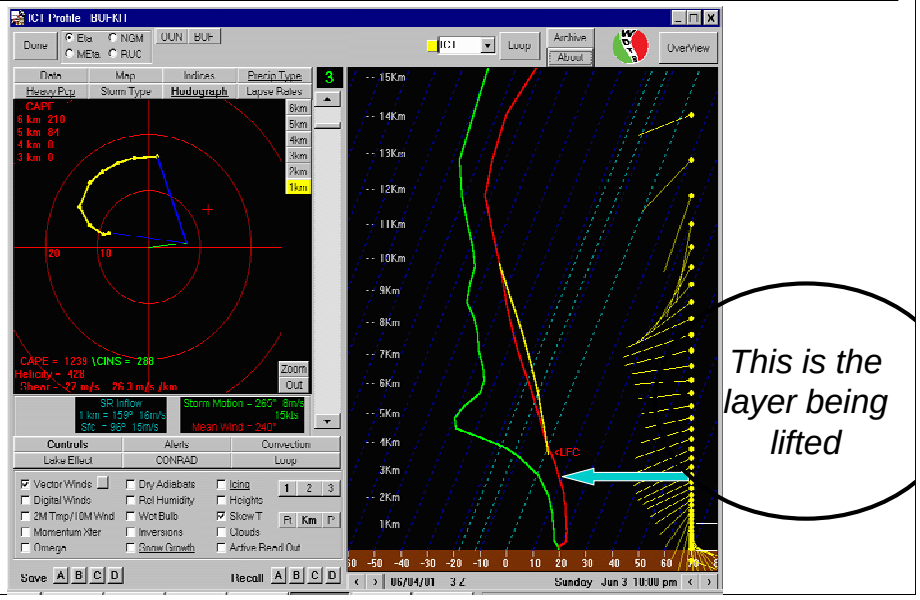
Details of Lift

- Now, let's look at a sounding at Point C



This sounding would be back in the colder air.

Elevated Convection
(sample sounding at Point C)

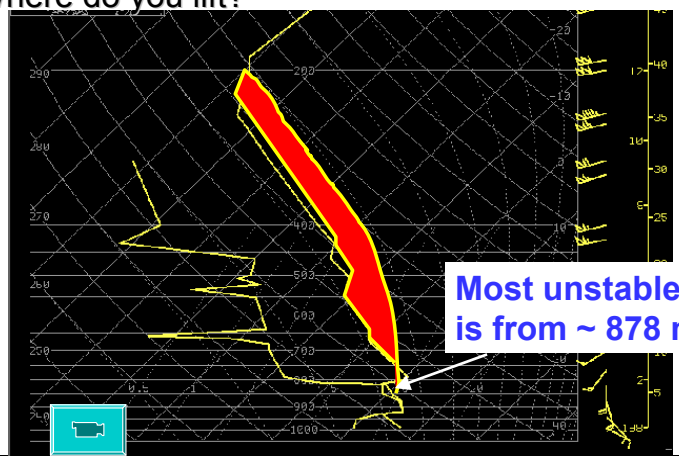


Look at lifting forces in the 2-4 km layer (6.5 – 13.1 kft) (800-625 mb), not surface to 850mb. That's why Warm Air advection centered around 850 mb is often a good correlation to lifting forces for elevated severe convection.

Important to assess convergence (theta-E convergence in 850-700 mb) and boundary relative flow. Also, frontogenesis in a 50-100 mb layer below the bottom level of positive buoyancy.

Details of Lift

- Consider this sounding…
  - How much lift is necessary for convection?
  - Where do you lift?



If you lift the parcel from 878 mb ,you get the most CAPE. This sounding was a proximity sounding for the big hailstorm that hit Dallas-Ft. Worth area.

For elevated convection , use the most unstable parcel (MUCAPE in SPC plots). Be aware of the limitations; it can overestimate the degree of instability and erode the cap too early before convection develops. Sometimes, it is better to use a lower mean layer for surface based convection.

There is an animation showing the effects of adjusting lifting parcel level with this sounding.

Orientation of Lift

- There are some additional factors to consider when evaluating how strong (and deep) the lift will be:
  - Boundary-relative flow (low-levels and steering layer)
  - Boundary-orthogonal shear (RKW Theory)

In addition to evaluating the individual details of parcels for lifting, we should look at the characteristics of lift associated with boundaries in terms of how strong and deep the lifting will be. Based on observations by Wilson and Megenhardt (1997), boundaries in FL were most active if cell motion -boundary motion was $\lesssim$ 5 m/s. Before storms, the 2 - 4 km mean wind = cell motion. Several observational studies including this one sampled the mean flow in a 2-4 km layer and found the boundaries were most convectively active when the boundary-normal flow was less than 5 m/s. The 2-4 km layer was chosen since it correlated best with new storm motion. Boundary-relative flows greater than 5 m/s exhibited less storm activity and those storms that did form on the boundary would quickly be left behind in less stable air and dissipate.

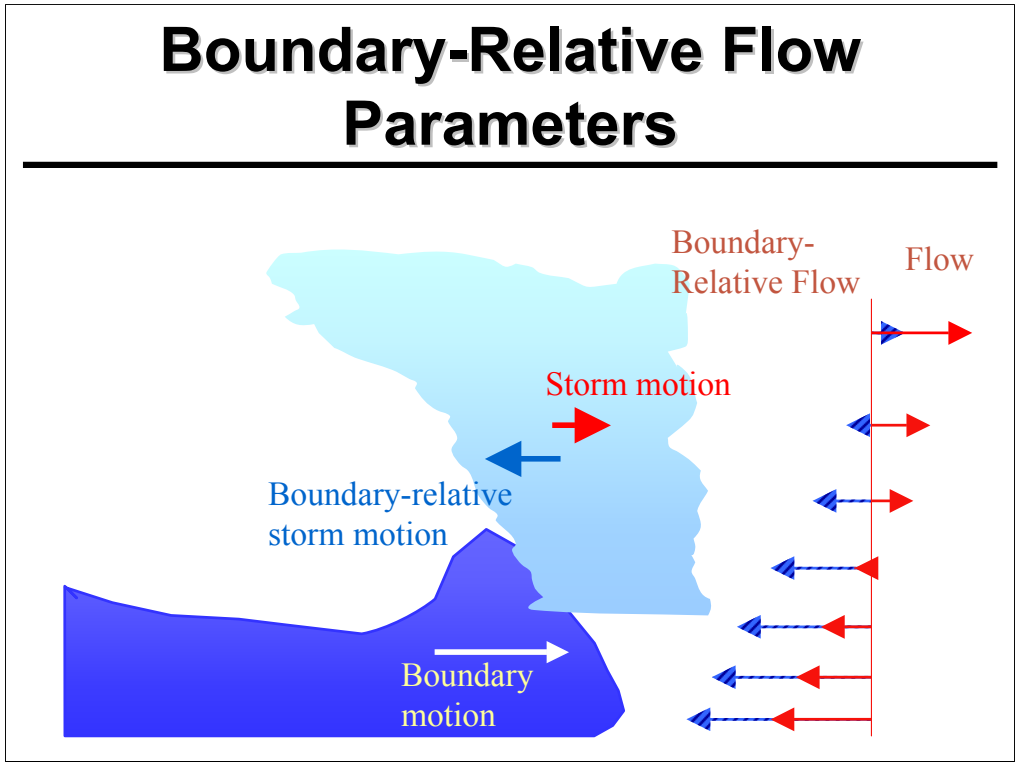
Note: This study was done during a field project around Cape Kennedy during the summer. Therefore the storms were mostly driven by surface-based phenomenon and there were small amounts of shear. Whether these findings can apply to stronger synoptically forced situations still remains a question worthy of testing at any of your locations.

Also note that not all storms will dissipate if the boundary-relative 2-4 km flow is high. Convection can persist or even thrive well after they leave the original boundary if there is still sufficient uncapped instability.

Boundary-Relative Flow Considerations

- Weak (0-5 m/s) boundary-relative (steering-layer) flow increases convective cell coverage
- Orientation of line of forcing can be as important as CAPE and shear in determining type of outbreak
  - Orientation WRT shear & wind vector
  - Boundary-relative supercell motion

Results of simulations (Bluestein and Weisman, 2000) suggest that when the deep-layer shear is normal to the line of forcing, isolated, cyclonically (anticyclonically) rotating, right-moving (left-moving) supercells develop at the end of the line to the right (left) of the deep-layer shear. A squall line having embedded, but distinct, intense cells develops in between. The squall line develops after neighboring right- and left-moving cells collide. The cells in the squall line have some supercellular characteristics, but the updrafts at midlevels are not very well correlated with vertical vorticity.

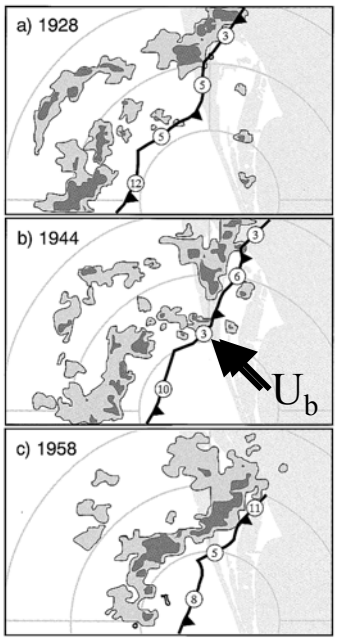


Graphic courtesy of Jim LaDue. These are theories that apply to density current boundaries:

1) A boundary-relative headwind (tailwind) increases (decreases) the depth of the ascending air column.
2) Based on numerical modeling of an outflow boundary, the boundary height increases when a boundary moves into a headwind.
3) Boundaries with a headwind move more slowly than those with a tailwind (from Moncrieff and Liu, 1999).

From these theoretical and numerical experiments, the ascending air column will be deeper for boundaries experiencing a significant headwind than for those following into a tailwind. It stands to reason that low-level convergence with headwind (tailwind) boundaries is higher (lower).

Boundary-Relative Flow



- Wilson and Megenhardt found the most coverage with boundary-relative storm motion (U_b) less than ± 5 m/s.

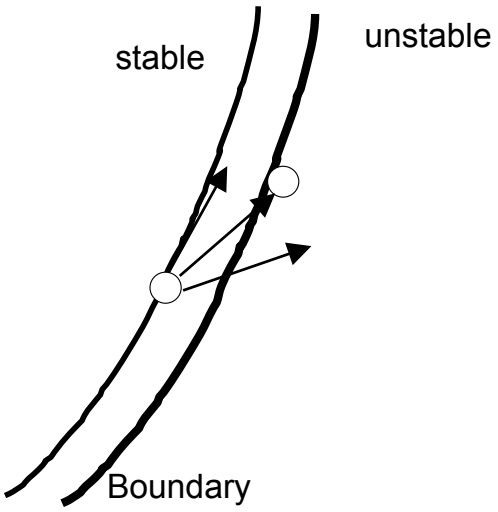
Wilson and Megenhardt (1997)

Wilson and Megenhardt (1997) found the most coverage with boundary-relative storm motion (U_b) less than ± 5 m/s.

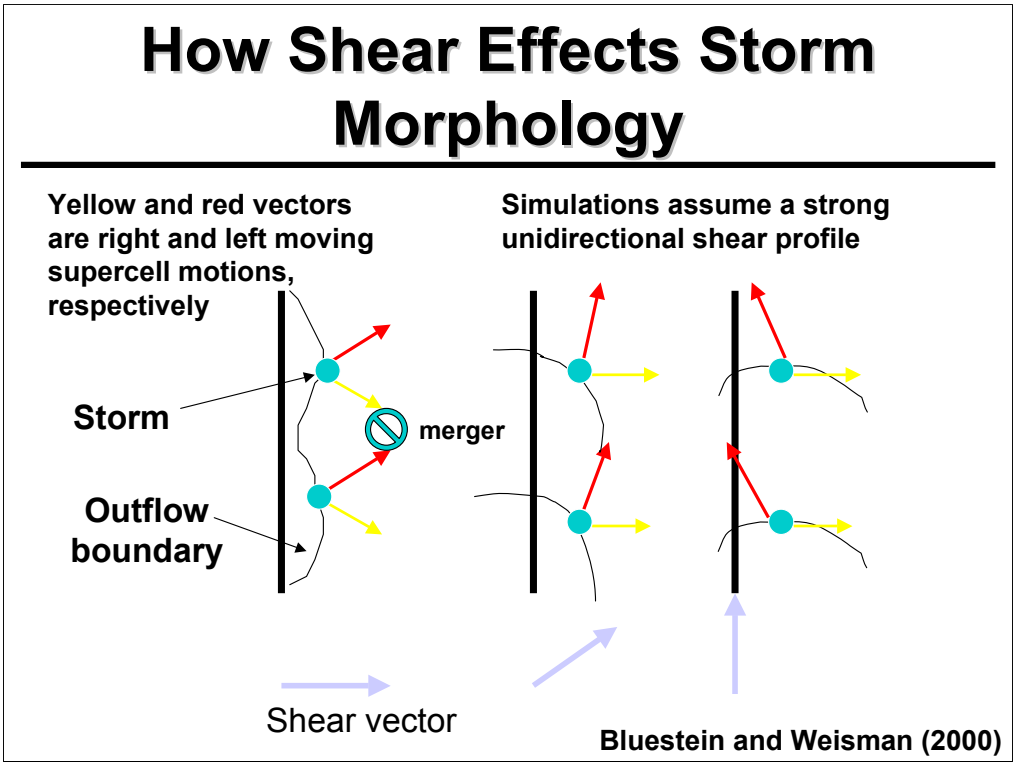
These storms occurred in weakly sheared environments.

Effects of Boundary-Relative Kinematics on Storm Morphology

- Steering flow
  - Is the storm going to remain on, fall behind, or overtake a boundary?
  - This may affect storm type in addition to CAPE and shear

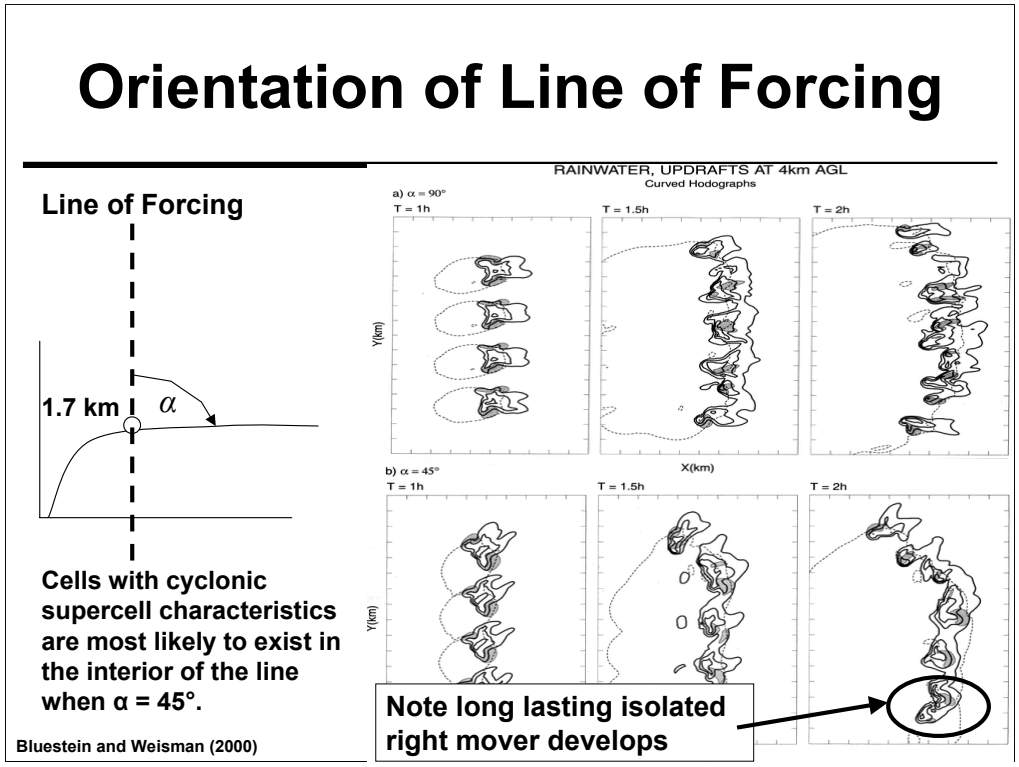


Based on boundary-relative considerations, the northward moving storm has the lowest boundary-relative flow which allows parcels to ride along the boundary and reach its LFC the fastest, thus maximizing lifting potential.



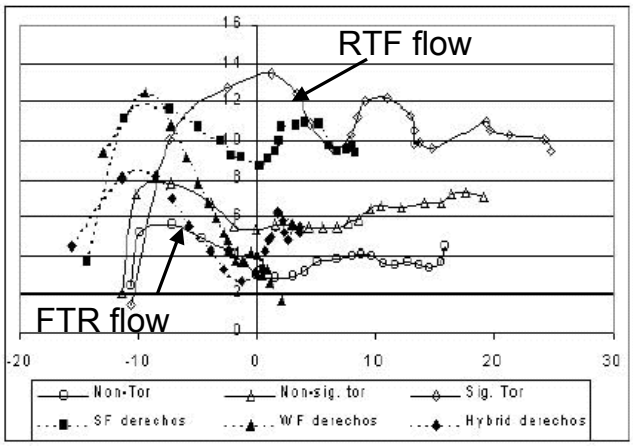
This is based on numerical simulation of storms (Bluestein and Weisman, 2000) where they looked at various orientations of the vertical shear vector with respect to the lines of forcing. The left-most example is of a shear vector oriented at a 90 deg. angle to the boundary. In this case, the right-mover will collide with the left-mover. In the middle situation, the shear vector is at a 45 deg. angle to the boundary. With the exception of the northernmost storm, the left-moving members will move across the outflow bndry generated by the right-moving neighbor to the north, and weaken, while the right movers will move away from the boundary and may not collide with their neighbors. The last case (far right) is where the shear vector is oriented parallel to the boundary. In this scenario, the left movers will move behind the boundary, while the right movers will slowly move ahead of the boundary. However, since there is also a component of storm motion parallel to bndry, the right movers might be able to catch up to the outflow bndrys generated by the adjacent right movers along the line, and weaken.

Thus, boundary-orthogonal shear (component of the shear vector normal to the boundary orientation) affects the motion of storms and thus, how long updrafts will stay on the boundary. The shear vector is drawn perpendicular to the gust front (thin lines), with right (yellow vectors) and left (red vectors) supercell storm motions based on Bunker's et al. (1999) method.



From Bluestein and Weisman (2000) , simulated fields for a line simulation, at 1, 1.5, and 2 h, of liquid-water mixing ratio (thick contours at 1, 4, and 8 g kg^{-1}) and vertical velocity (shaded region in excess of 8 m s^{-1}) at 4 km for the modified sounding and curved hodograph. Tick marks are spaced 10 km apart. Gust front at 0.25 km depicted by dashed line (perturbation temperature at 0.25 km, with respect to ambient environment, of −1°C). This simulation was for when alpha (angle between shear vector above 1.7 km AGL and line of forcing) was = 90 deg. In reality,this would be for a westerly shear and a north-south boundary.

Storm-Relative Winds Affect Mode of Convection

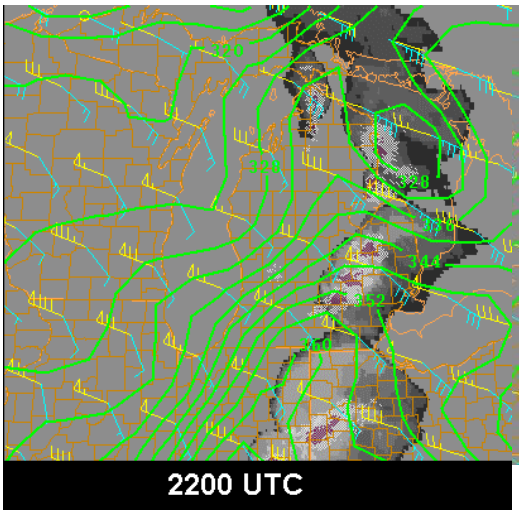


From Evans and Doswell (2002)

When storm-relative winds are examined (as in this figure from Evans and Doswell, 2002) it is apparent that the derechos yield the strongest inflow in the lowest 1 km. In addition, weakly forced events develop and persist in environments with deep storm relative inflow (front-to-rear flow) from the surface through 8-9 km. In contrast, the supercell dataset reveals pronounced rear-to-front flow above 2-3 km , especially the significantly tornadic events. This is really evident above 4 km, where only the supercells indicate rear-to-front storm-relative flow increasing through 10 km. These results were consistent with other studies that found the distribution of hydrometers and precipitation largely due to the mid- and upper-level wind fields *relative* to storms.

Boundary-Relative Flow
7/2/97– Michigan Tornadoes

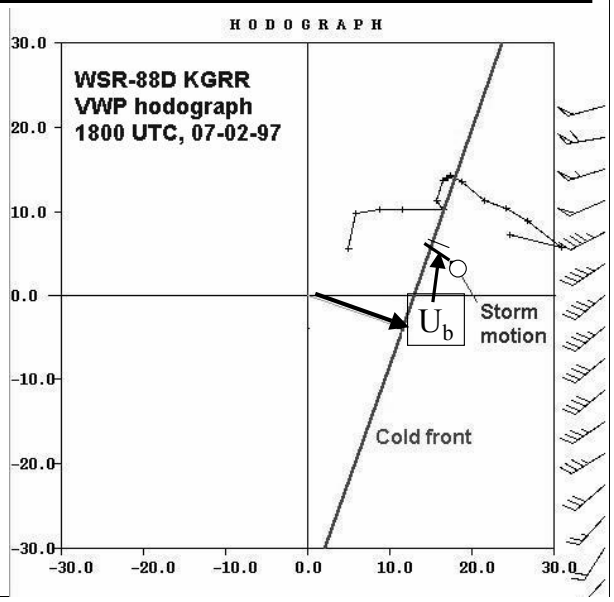
- Strong synoptic forcing and favorable CAPE/shear for tornadic supercells
- Boundary-relative flow in mid to upper levels was rear-to-front



The blue vectors are 850 mb boundary orthogonal winds, the yellow vectors are anvil level (~ 250) boundary orthogonal winds (after LaDue, 1998).

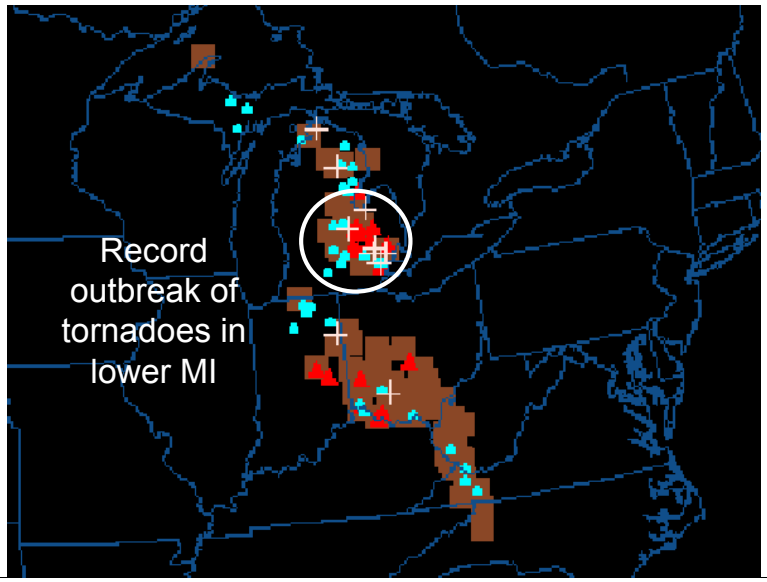
7/2/97 Michigan Tornadoes

- VWP hodograph shows observed storm motion pulling ahead of front



The blue dot is the storm motion vector. Boundary relative flow is the blue vector (~ 280 deg, 5 m/s).

7/2/97 Storm Summary

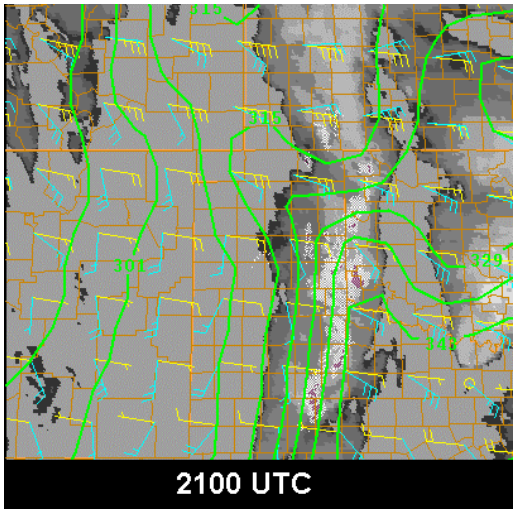


Severe weather reports from 1200 to 2300 UCT 02 July 1997. Small blue square-like symbols are hail >= 0.75 inches in diameter, red triangles are tornadoes, large brown squares are wind damage, white crosses are reported wind gusts >= 50 knots.

See http://www.wdtb.noaa.gov/resources/cases/jul02case/torcase.htm for more information on this event.

May 5, 1995 – TX PH Tornado Null Case

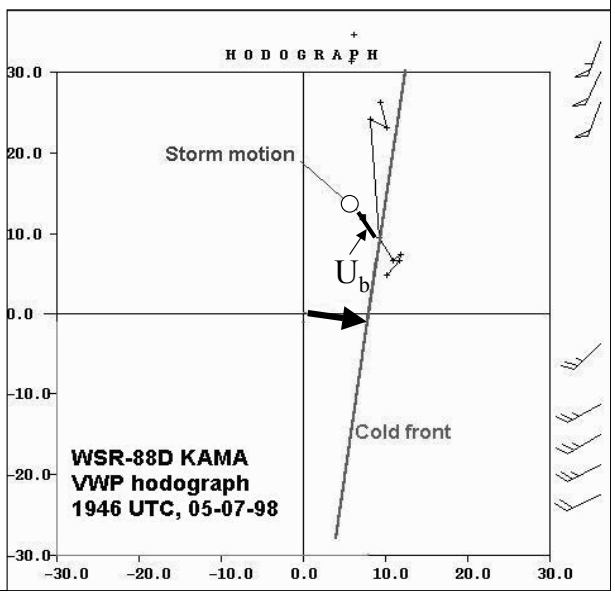
- As in 7/2/97 case, shear and buoyancy were in range of supercell tornadoes
- However, boundary-relative upper level flow was front-to-rear



Same vector notation as previous example.

May 05, 1995 – TX PH tornado null case

- VWP hodograph shows observed storm motion falling behind front



Same color convention as previous example.

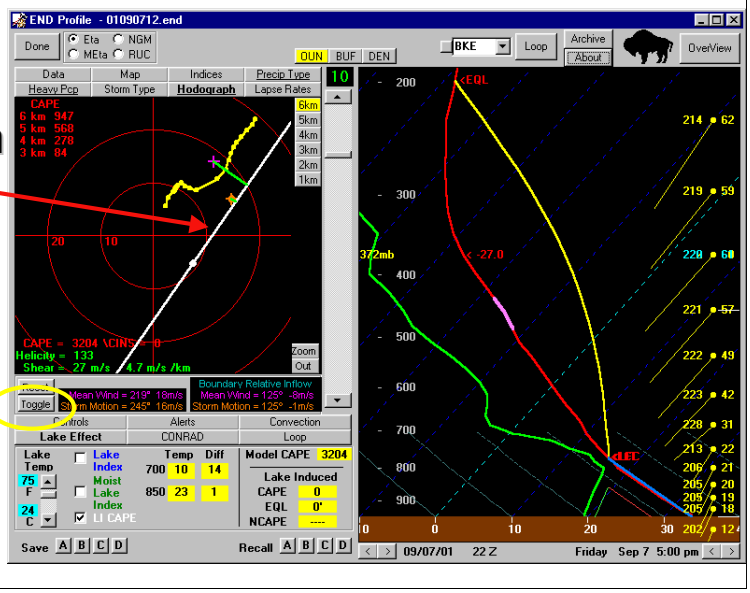
Visualizing Boundary-Relative Flow/Shear

- Line of forcing can be viewed as another axis on a hodograph
- Winds and shear can be interpreted with respect to the line of forcing
- The line of forcing can be displayed with the BUFKIT hodograph

The best way to view orientations of line of forcing and shear is through a hodograph like what is available on BUFKIT.

Visualizing in BUFKIT

- Boundary Orientation

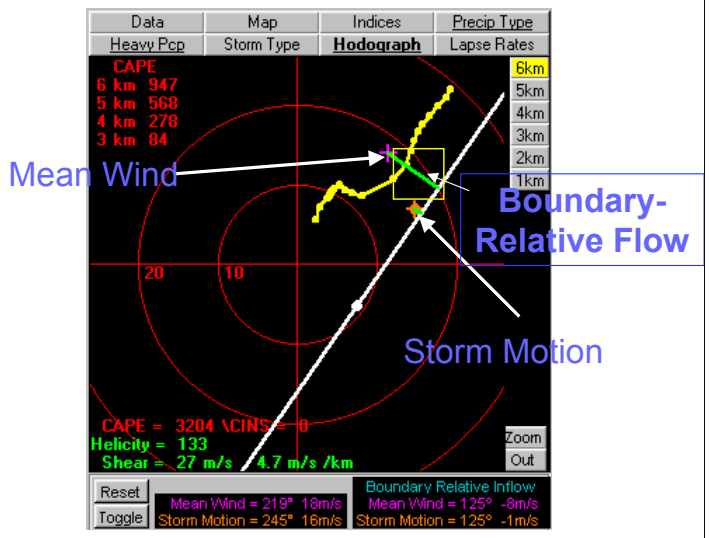


Use the toggle button to go from hodograph display to B-R coordinates

This hodograph is from a Eta forecast profile at Enid, OK.

Visualizing in BUFKIT

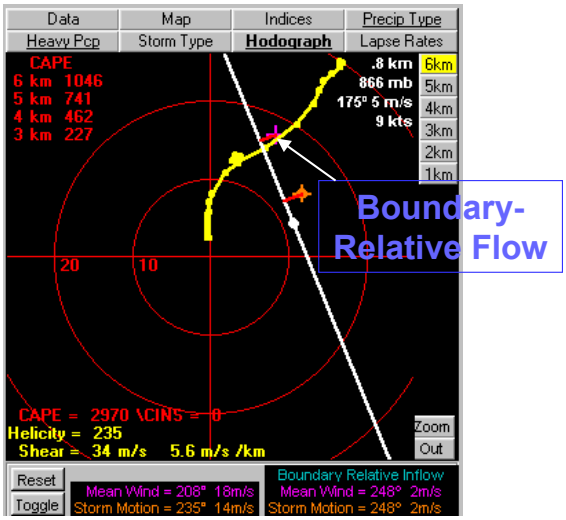
- 10 hr Eta forecast for END
- Storms forecast to fall behind boundary



This shows storms falling behind the boundary due to BRF considerations. On BUFKIT, the green vector is the Boundary-relative flow vector.

Visualizing in BUFKIT

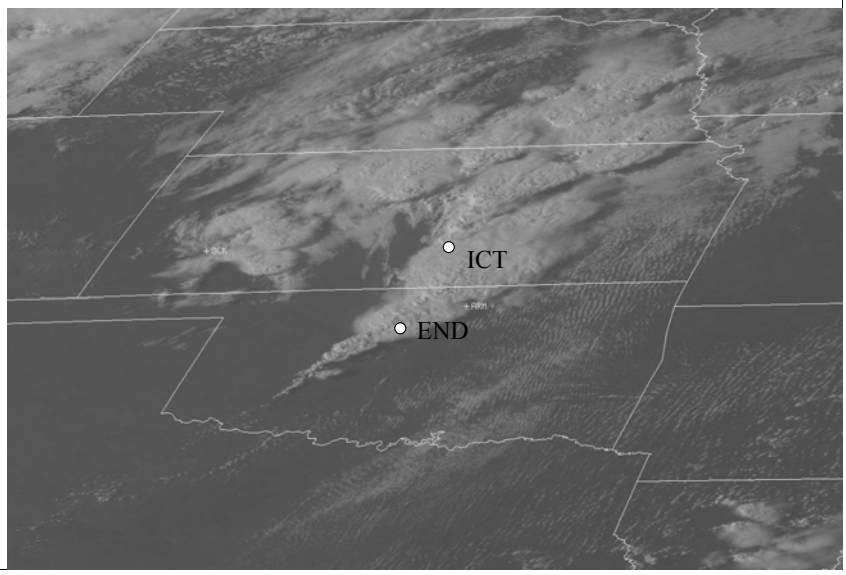
- 10 hr Eta forecast for ICT
- Storms forecast to stay just ahead of boundary



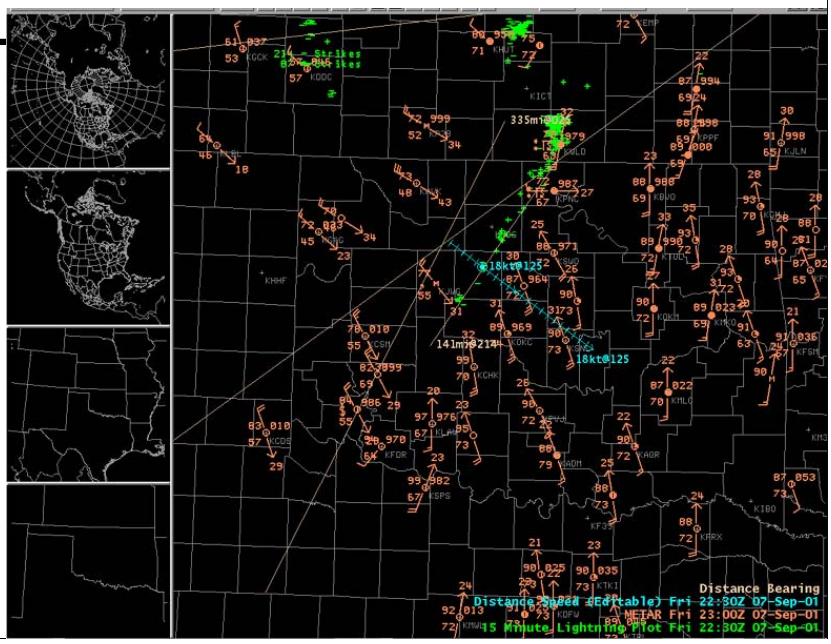
This forecast shows storms staying out ahead of the boundary in the warmer air.

Sept 07, 2001 case

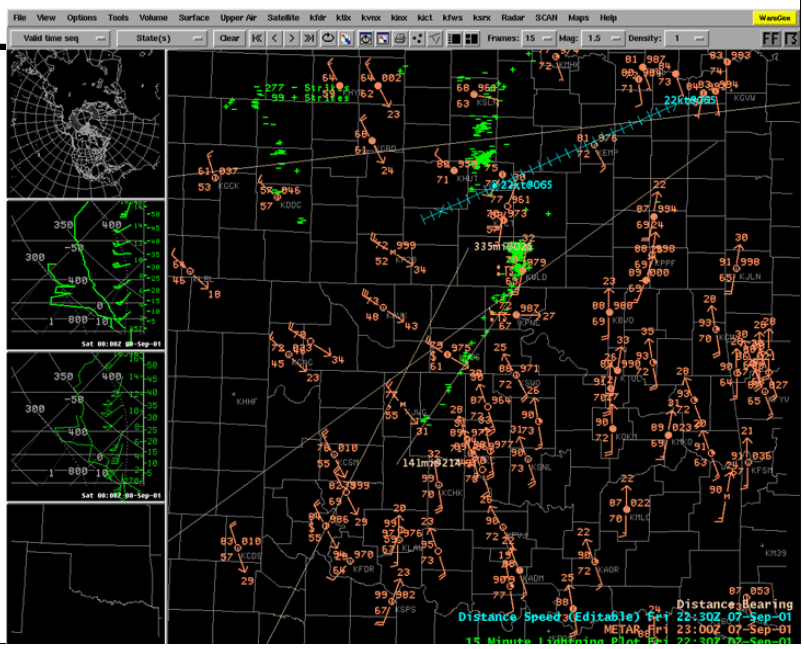
GOES-8
2200 UTC



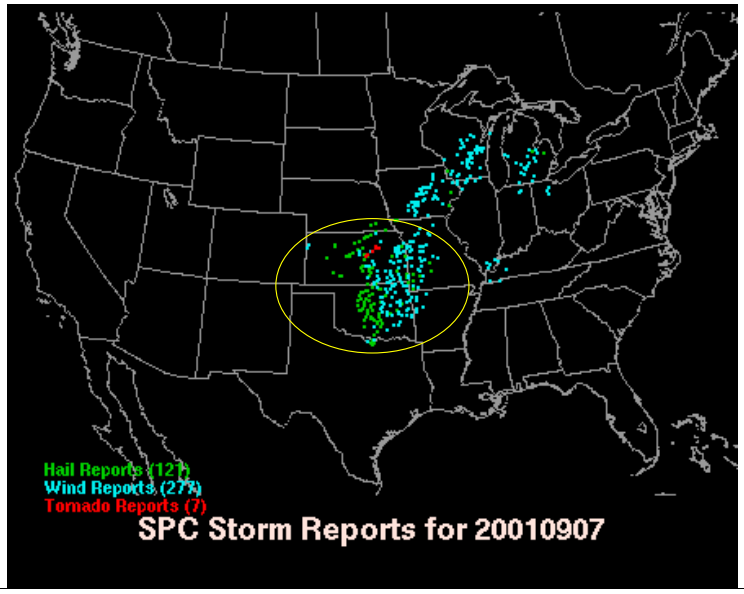
Sept 07, 2001 case



Sept 07, 2001 case



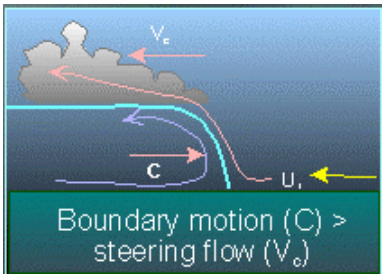
Sept 07, 2001 case



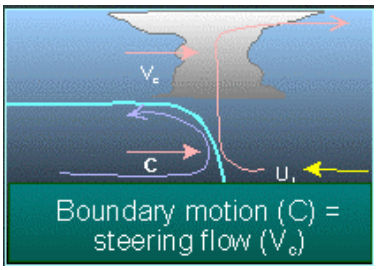
Boundary orientation relative to mean wind and shear appeared to influence convective mode, another consideration to lifting characteristics and your threat assessment.

Depth of Ascent Zone

- Key is to maximize residence time of parcel in ascent zone to ensure parcel reaches the LFC
  - Related to deep layer shear but also could be any boundary
  - Small (≤ 5 m/s) boundary-relative motion



Destructive to lifting



Promotes deeper lifting

This theory is based on Rotunno, Klemp, and Weisman (RKW) numerical simulation studies from the 80s and early 90s. It is based on horizontal vorticity arguments. The key to maximizing residence time of a parcel in the ascent zone is to ensure it reaches the LFC. (See http://meted.ucar.edu/mesoprim/shear/frameset.htm). These theories apply to density current boundaries. Density current boundaries include strong fronts, seabreeze boundaries and outflow boundaries. Deepest ascent occurs when boundary orthogonal environmental shear balances cold pool vorticity (Rotunno et al. 1988), or when the boundary speed (C) cancels out positive shear magnitude (U_L). C/ U_L ~ 1. When the boundary-normal (or orthogonal) component of the shear is oriented forward away from the cold side of the boundary (positive shear) and it has the same magnitude as the forward speed of the boundary, the conditions are optimal for the strongest, most vertically oriented ascent zone. That is C/ Delta U_L ~ 1. This is graphically shown in the right side schematic. What often occurs is shown in the left side figure where the shear vector is directed front to rear of the boundary (negative shear). In that case, no balance occurs in C/ UL and the ascent zone becomes much shallower. When a boundary does not have any density current characteristics, the shear considerations do not apply in the sense in which they were presented. In some sense shear is still important since boundary-normal shear can be too strong potentially ripping apart an ascending zone of air before it can reach the LFC.

Determining Depth of Ascent Zone – 3 main factors

1. Maximize low-level headwinds
   - Applies to density current boundaries
   - Not necessarily strong convergence
2. Boundary-orthogonal shear
   - For isolated storms (eg. supercells), like to see rear-to-front flow from 2-3 km through 10 km equal to boundary motion
3. Boundary-orthogonal mean flow
   - For isolated storms, best if this value is $\leq$ 5 m/s

This is a summary of topics covered related to determining the ascent zone.

Summary
Lesson 2

- Factors in assessing lift
  - Boundary-relative flow, shear/cold pool relationships
- Types of lifting
  - Synoptic or mesoscale
- Orientation of lines of forcing
  - Can affect intensity and longevity of storms